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1 **Heterogeneity in Karakoram glacier surges**

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8

9 **Abstract**

10 Many Karakoram glaciers periodically undergo surges, during which large volumes of ice and
11 debris are rapidly transported down-glacier, usually at a rate of one to two orders of magnitude
12 greater than during quiescence. Here we identify eight recent surges in the region, and map their
13 surface velocities using cross-correlation feature tracking on optical satellite imagery. In total, we
14 present 44 surface velocity datasets, which show that Karakoram surges are generally short-lived
15 (between 3 and 5 years in most cases), have rapid build-up and relaxation phases (often less than a
16 year), and generally reach peak velocity during summer months. Otherwise, they do not follow a
17 clearly identifiable pattern. In two of the surges, the peak velocity travels down-ice through time as
18 a wave, which we interpret as a surge front. Others are characterised by high velocities that occur
19 simultaneously across the entire glacier surface and acceleration and deceleration is close to
20 monotonic. There is also no consistent seasonal control on surge initiation or termination. We
21 suggest that the differing styles of surge can be accounted for by individual glacier geometries, and
22 that while some characteristics of Karakoram surges are akin to thermally-controlled surges
23 elsewhere (e.g. Svalbard), the dominant surge mechanism remains unclear. We thus propose that

24 classic thermal and hydrological surge classifications are not appropriate in the Karakoram context
25 because the surges cannot be collectively categorised. The implication of this is that regional
26 triggers may also not be singularly defined, and may even differ on a glacier by glacier basis.

27 **1. Introduction**

28 Glacier surges are reported from the Canadian and Russian High Arctic, Svalbard, Iceland,
29 Greenland, Alaska and parts of the Himalaya. These surge-type glaciers undergo cyclical non-
30 steady flow consisting of two distinct phases (Meier and Post, 1969). The active phase, typically
31 lasting a few months to a few years, is a period of activity during which glacier velocity increases
32 by at least an order of magnitude. The quiescent phase, typically lasting tens to a few hundreds of
33 years, is a period of relative stagnation during which the lower portion of the glacier (the receiving
34 area) thins, and mass builds up in an upper, reservoir area. During surges, mass is rapidly
35 transferred from the reservoir to the receiving area, and an advance of the glacier terminus often,
36 but not always, takes place.

37 Two ‘types’ of glacier surge (thermally-regulated and hydrologically controlled) have long been
38 referred to in the literature, which describe the trigger mechanisms by which an active phase is
39 initiated. In the first, changes in basal temperature promote increased sediment deformation and
40 porosity and a positive feedback between pore water pressure, deformation and basal flow ensues
41 (Clarke et al., 1984; Murray et al., 2000). These thermally regulated surges are characterised by
42 several years of acceleration before the peak of the surge is reached (often termed the initiation
43 phase), several years of deceleration following the peak of the surge (often termed the termination
44 phase), and tend to begin their acceleration/deceleration independent of any seasonal control. They
45 are mostly recognised in Svalbard (Murray et al., 2003) and the Yukon (Clarke et al., 1984). In the
46 second, changes in the efficiency of the hydrological system (and thus pore water pressure) trigger
47 the flow instability (Kamb et al., 1985; Björnsson, 1998). Such hydrologically regulated surges are
48 characterised by rapid acceleration and deceleration (i.e. days to weeks long), and tend to initiate

49 during winter months (a time of drainage inefficiency) and terminate during summer months (when
50 efficiency is increased). Such events are mostly recognised in Alaska (Burgess et al., 2012; Lingle
51 and Fatland, 2003).

52 Remotely sensed data have provided the foundation for many contemporary studies of surge-
53 type behaviour (e.g. Fatland and Lingle, 1998; Murray et al., 2003; Quincey et al., 2011; Mayer et
54 al., 2011; Turrin et al., 2013). Velocity data are derived using either cross-correlation feature
55 tracking (of either optical imagery or synthetic aperture radar imagery, or both) or interferometry
56 (where the surge is slow enough to maintain coherence), and studies have focussed on rates of
57 kinematic wave propagation (Turrin et al., 2013), surge return periods (Quincey and Luckman,
58 2014) and the contribution of surging glaciers to tidewater ice fluxes (Burgess et al., 2014). Many
59 studies have focussed on identifying trigger mechanisms (e.g. Murray et al., 2003), but for some
60 regions of the world the mechanics of glacier surging remain poorly understood. This is particularly
61 true in remote terrain, where surges may go entirely undetected or only be recognised once
62 underway. One such region is the Karakoram, Pakistan, which is home to one of the largest
63 concentrations of surging glaciers anywhere in the world (Copland et al., 2011), but remains
64 inaccessible for many researchers because of ongoing political tension.

65 Better quantification of glacier surge dynamics (magnitude of, and spatial variability in
66 acceleration and deceleration) and how they differ within and between regions are important to
67 realise if the basal processes that yield such rapid changes are to be understood. In high-elevation
68 regions such as the Karakoram, this also has important implications for landscape evolution (in
69 terms of erosion/deposition) as well as local water supplies and hazard development (in terms of
70 land inundation, ice/rock avalanching from surging masses, and ice-dammed lake development).
71 Thus, the aim of this paper is to augment the limited surge data we have for the region already
72 (Quincey et al., 2011; Mayer et al., 2011) with measured changes in surface velocity on eight
73 further Karakoram valley glaciers during recent surge events. These new results indicate there are
74 distinct similarities between Karakoram surges and those documented in Svalbard (e.g. Murray et

75 al., 2003), but that some dynamic characteristics are more consistent with a hydrological control, as
76 has been suggested elsewhere (Mayer et al., 2011). We conclude that Karakoram surges do not fit
77 neatly within long-standing dynamic models of surge-type behaviour, and suggest that unstable
78 flow should be viewed as a continuum rather than as a binary classification.

79 **2. Study area: Karakoram glaciers**

80 The majority of glaciers in the Himalaya are receding and have lost significant mass since at
81 least 1970, despite thick debris cover (Bolch et al., 2011; Kääb et al., 2014). Glacier wastage is
82 spatially heterogeneous, and is linked to both topography and climate (Fujita and Nuimura, 2011).
83 More than 65% of the monsoon-influenced glaciers studied by Scherler et al. (2011) were observed
84 to be receding. However, heavily debris-covered Karakoram glaciers with stagnant low-gradient
85 terminus regions typically have stable fronts and indeed some glaciers are advancing as increased
86 surface debris cover retards glacier melt (Scherler et al., 2011). Other Karakoram glaciers advance
87 periodically during surges, when velocities increase rapidly to rates between one and two orders of
88 magnitude greater than during quiescence (Hewitt, 1969). Previous work has suggested a
89 preponderance of surge-type behaviour in glaciers between 12 and 25 km in length (Hewitt, 1969)
90 and those fed by tributary glaciers (Hewitt, 2007). The season of Karakoram glacier surge initiation
91 varies, and surges have been shown to develop gradually over several years (Quincey et al., 2011).
92 These can lead to km-scale advances of glacier termini over very short (monthly to annual)
93 timescales.

94 Previous work focussing on the triggers of Karakoram surges have arrived at conflicting
95 conclusions (Quincey et al., 2011; Mayer et al., 2011). On one hand, Karakoram glacier surges have
96 been suggested to be thermally rather than hydrologically controlled, coinciding with high-altitude
97 warming from long-term precipitation and accumulation patterns (Quincey et al., 2011; Quincey et
98 al., 2014). On the other, observations and modelling from a single surge event invoked a change in
99 hydrological conditions as the main trigger mechanism (Mayer et al., 2011). There is some

100 consensus that glacier surges are increasing in frequency in the region, but return periods are poorly
101 constrained. Estimates and observations normally cite typical return periods of the order of 25-40
102 years (Guo et al., 2013; Copland et al., 2011), although historical observations of the Khurdopin
103 Glacier suggest a slightly shorter return period of ~20 years (Mason, 1930; Quincey and Luckman,
104 2014).

105 Here we present data on glacier velocity and changes in the surface character of eight
106 Karakoram glaciers through recent surges (Figure 1; Table 1). The glaciers vary in character from
107 long, debris-covered tongues, the longest of which is the Skamri Glacier (~40 km) located directly
108 to the east of the Shimshal Valley, to short (i.e. < 15 km length), debris-free glaciers that are
109 unnamed, at least in the scientific literature (Figure 2). Five of the glaciers are already known to be
110 surge-type (e.g. Braldu, Chong Khumdan, West Qogori, Skamri and Saxinitulu; Copland et al.,
111 2011; Gardelle et al., 2012), while the other three have not previously been identified as surge-type.
112 The contrasting dynamics of the eight surges combined with their distinct surface geomorphologies
113 provides the opportunity to evaluate the processes controlling surge initiation and development in
114 more detail than has previously been possible.

115 **3. Methods**

116 Multi-temporal velocity fields were calculated by cross-correlation feature-tracking (Strozzi et
117 al., 2002). This method has been repeatedly shown to produce high-quality results on Himalayan
118 and Karakoram glaciers because of the abundance of surface features associated with debris-cover
119 and surge-type flow (Quincey et al., 2009; Mayer et al., 2011; Quincey et al., 2011). Satellite
120 images were sourced from Landsat TM, Landsat ETM+, Landsat OLI, ALOS AVNIR and ASTER
121 sensors (Table 2) to give as dense a dataset as possible through each of the surges. The feature-
122 tracking approach has been well-described elsewhere so we provide a summary of our approach
123 here. In the case of the AVNIR and ASTER data, the first step was to orthorectify the images using
124 the automated function (based on sensor model and digital elevation data) within ENVI 5.1. All

125 Landsat imagery was provided at L3, with the orthorectification already carried out by USGS. The
126 images were then co-registered on an individual glacier scale to correct for remaining misalignment.
127 We used coarse windows of 128 x 128 (pattern size) and 256 x 256 (search area) to achieve this.
128 Horizontal ground displacements were extracted using a Fourier-based correlation technique
129 (Luckman et al., 2007) with search windows of between 24 x 24 to 64 x 64 pixels (pattern size), and
130 32 x 32 to 128 x 128 pixels (search area).

131 Errors in the resulting displacement data arise from mis-registration of the two satellite images
132 and the precision of the algorithm used. Our co-registration is sub-pixel, and is therefore likely to be
133 similar to the ~5 m accuracy quoted by Lee et al. (2004) when considering multi-temporal Landsat
134 7 ETM+ images acquired on the same path and row. The correlation technique is affected by
135 changes in crevasses and surface debris patterns through time and space as well as the potential for
136 mis-matches of surface features. To mitigate against the latter errors, resultant displacement data
137 were filtered using signal-to-noise ratio as the primary indicator of the quality of the match. We also
138 removed extreme values (i.e. above a stipulated max threshold) and removed matches that did not
139 conform to the general flow direction of the glacier, defined manually by the user. This left only the
140 most robust patch correlations, for which the measurements themselves are expected to be of sub-
141 pixel accuracy.

142 To provide an indication of the uncertainty (σ) in the remaining velocity values we used the
143 following equation, modified from McNabb et al. (2012):

$$\sigma = 365 \frac{C_{pix} C_{match} \Delta x}{\Delta t}$$

144 where C_{pix} is the uncertainty in co-registration in pixels, C_{match} is the uncertainty in the matching
145 algorithm, Δx is the image resolution in metres, and Δt is the time interval between the image pair in
146 days. The highest uncertainty is thus associated with short (16-day) data separations (Table 2).

147 However, it should be noted that as these data coincide with the peak surge velocities, the measured
148 displacements still far exceed the potential errors.

149 To aid interpretation of the surge dynamics, surface debris structures were mapped for every
150 glacier using time-separated optical satellite images in ArcGIS. Features mapped include glacier
151 extent, areas of surface debris and associated surface debris structure.

152 **4. Results**

153 Forty-four velocity fields were derived through the eight glacier surges (Figure 3). It should be
154 noted that our derived velocity data are generally restricted to the ablation area, so our analysis does
155 not focus on dynamics in the accumulation zone. Centreline profiles show the magnitude and timing
156 of each event as it impacts the lower part of the glacier (Figure 4; n.b. we do not plot error bars here
157 to avoid obscuring data patterns). The maximum velocity recorded in any of the datasets was $\sim 2 \text{ km}$
158 a^{-1} and in all cases the peak surge velocities exceeded those in the build-up period by at least one,
159 and in some cases two, orders of magnitude.

160 While it is difficult to identify exactly when each of the surges initiated, some insight can be
161 drawn from looking at the differences between individual profiles. In the case of the first unnamed
162 glacier (Unnamed1), there was relatively slow flow during the summer months of 2009, but the
163 surge was fully developed by the early summer months of 2010, indicating that sometime during the
164 winter months of 2009 the switch between slow and fast flow took place. Similarly, the Shakesiga
165 surge was in its infancy during the late summer of 2009, but had reached its maximum velocity by
166 mid-summer of 2010, again indicating the switch took place during winter months. In the case of
167 the second unnamed glacier (Unnamed2), the surge appears to have been developing during the
168 summer months of 2006 and actually receded during the following winter months before switching
169 to fast flow in the summer of 2007. The initiation phase is missing in the available data for several
170 of the other surges, but the data from the Skamri Glacier also suggest that the switch to fast flow

171 took place more towards the summer season than the winter. In all cases it appears that the initiation
172 phase was months to years long.

173 The termination periods also appear to have been variable in their timing. Perhaps the best
174 defined is that of Unnamed1, where the surge was clearly active during the summer months of 2010
175 but began decelerating at the start of the following winter (in the November dataset). The Shakesiga
176 surge follows a similar dynamic, with the surge appearing to diminish in the early winter of 2010
177 having peaked in the immediately preceding summer months. In several other cases the termination
178 phase was slow to develop, and thus identifying when the switch from fast to slow flow took place
179 becomes difficult. Nevertheless, it appears that the termination phase was longer than the initiation
180 phase in the datasets where observations for both are possible (four of the eight datasets –
181 Shakesiga, Unnamed1, Unnamed2 and Skamri). In all four of these cases, the total surge lasted for
182 between 3 and 5 years; in a fifth (Saxinitulu) the surge is still ongoing, eight years after initiation.

183 In common with previous observations on the Kunyang Glacier (Quincey et al., 2011), at least
184 two of the currently studied glacier surges are characterised by a down-glacier propagation of the
185 velocity peak. We interpret this to represent the surge front, although we have no surface elevation
186 data to confirm its topographic expression. In the case of the Braldu surge, there is a clear velocity
187 wave that propagates down-glacier at approximately 2 km a^{-1} at the height of the surge (Figure 4).
188 There is a less-clear front in the Unnamed1 dataset, but during the summer of 2010 the peak
189 velocity did migrate down-glacier and its arrival at the glacier terminus coincided with a
190 deceleration both around the terminus and up-glacier. There are also hints of a surge front in both
191 the Chong Khumdan and Skamri datasets, but based only on limited data. In contrast, other glaciers
192 show a very different dynamic, with the surge affecting almost the whole glacier coincidentally.
193 The Shakesiga dataset shows this most clearly, with a uniform increase in flow across the entire
194 glacier length. A similar, but less pronounced, increase is also visible in the Saxinitulu and Qiaogeli
195 surges. The Unnamed1 dataset shows characteristics of both surge styles, with a generally

196 monotonic acceleration/deceleration affecting the lowermost ~7 km of ice, but also showing some
197 evidence of a surge front.

198 Several of the smaller (<15 km) glaciers experienced major frontal advances (Figure 5) whereas
199 surges within the larger (>10 km) glaciers were mostly confined to the existing glacier area. The
200 Braldu surge, although still technically ongoing, does not look likely to impact the lowermost 10
201 km of debris-covered ice. Similarly, the Skamri surge looks to have terminated approximately 10
202 km from the terminus. The Shakesiga surge resulted in a small frontal advance of several hundred
203 metres, but not sufficient to override the main valley river and abut the opposing valley wall. Both
204 of the unnamed glaciers as well as the Saxinitulu Glacier and the Qiaogeli Glacier advanced by
205 several kilometres during their surges; indeed the Saxinitulu Glacier is still advancing at ~100 m per
206 year having already advanced 1 km from its original terminus position.

207 **5. Discussion**

208 Previous studies focussing on Karakoram surges have suggested both thermal and hydrological
209 controls may be responsible for their initiation (Quincey et al., 2011; Mayer et al., 2011). Evidence
210 that has supported the thermal switch hypothesis includes the apparently random timing of the
211 initiation phase and its length (usually several years, as opposed to the < 0.5 years observed in other
212 regions (Kamb et al., 1985)), as well as a surge-front identified in one dataset (Kunyang Glacier;
213 Quincey et al., 2014) that may have represented the boundary between the thawed and frozen bed
214 (cf. Fowler et al., 2001). Numerical modelling has been used to explain the propagation of a similar
215 surge front on the Gasherbrum Glacier using concepts of glacier sliding with cavitation and
216 subglacial hydrological switching, and to explain modulation waves (small amplitude velocity
217 peaks) identified in the feature-tracked velocity data (Mayer et al., 2011). Coupled with these
218 previous observations, multi-temporal velocity data now exist for twelve Karakoram surges
219 (Figures 4 and 6), including one duplicate, Khurdopin Glacier (Quincey et al., 2011; Quincey and
220 Luckman, 2014)). These combined data suggest that at least two types of surge exist in the
221 Karakoram: the first is characterised by a peak-velocity wave (which we interpret as a surge front)

222 propagating down-glacier; the second is characterised by more uniform and simultaneous
223 acceleration over the full glacier length.

224 Mayer et al. (2011) identified a surge front in their Gasherbrum velocity data, and Quincey et al.
225 (2014) reported similar observations on the Kunyang Glacier. Travelling waves have been observed
226 during many previous glacier surges, and have been linked to both hydrological trigger (Kamb et
227 al., 1985; Fowler, 1987) as well as thermal trigger mechanisms (Fowler et al., 2001). In the case of
228 the former, the surge front is thought to represent the transition between an efficient tunnel drainage
229 system promoting flow by deformation (down-glacier) and an inefficient linked-cavity system
230 promoting flow by sliding (up-glacier). It has been suggested that there may be a seasonal signal to
231 hydrologically controlled surge front propagation (Turrin et al., 2013; Raymond, 1987), with
232 deceleration during summer months when subglacial channelization reduces water pressure, and
233 acceleration during contrasting (hydrologically inefficient) winter conditions. In the case of the
234 thermal switch theory, the boundary is thought to be between warm-ice (up-glacier) and cold-ice
235 (down-glacier). According to Clarke (1976), the cold ice is immobile and frozen to its bed during
236 quiescence. The critical element in terms of whether a surge initiates appears to be the thickness and
237 permeability of the underlying sediment layer (Fowler et al., 2001), and where there is no restriction
238 to flow at the margin, the surge front may be entirely absent.

239 The Braldu surge is relatively short-lived and given the temporal resolution of the observations
240 it is difficult to determine any seasonal signal (or lack of signal) in the propagation of its surge
241 front. However, the fact that its down-glacier progression is inhibited by immobile (and probably
242 cold) ice is clear to see in both the velocity data (Figure 4a) and in the geomorphological
243 interpretation, which illustrates a long, stagnant, debris covered tongue (Figure 7). The other dataset
244 in which a surge-front may be present is Unnamed1. This glacier is particularly interesting because
245 the surge appears to have overridden debris (or even dead-ice) that is a remnant of a previous
246 advanced glacier position (Figure 8). In both cases, therefore, significant obstacles impeded the
247 surge. The same is true for the Kunyang surge identified in Quincey et al. (2014); the Kunyang

248 Glacier showed extensive areas of thermokarst pre-surge indicating stagnant or slow-moving ice,
249 and the main glacier into which the Kunyang feeds (Hispar Glacier) is known to be slow-flowing
250 (Rankl et al., 2014) and thus provides a further obstacle to fast-flowing ice. It is therefore possible
251 these surge fronts could simply be a consequence of the individual glacier geometries rather than
252 representing a thermal or drainage boundary as has been invoked elsewhere (Fowler et al., 2001;
253 Kamb et al., 1985).

254 In contrast, several of the gathered datasets show a much more uniform and spatially coincident
255 acceleration, akin to that observed at Monacobreen in Svalbard (Murray et al., 2003). The
256 equivalent end-member (in our Karakoram data) appears to be the Shakesiga dataset, although the
257 Saxinitulu and Qiaogeli surges and previous profiles for the Khurdopin Glacier and the Gasherbrum
258 Glacier (Figure 6) are similarly characterised. In such cases, the lack of a surge front could be
259 accounted for by a thermal activation front propagating faster than ice flow and consequently no
260 build-up of fast-flowing ice is apparent (Fowler et al., 2001). Similarly, the dynamic evolution of
261 surges observed on smaller glaciers in our dataset (Unnamed1, Unnamed2) also conform to
262 theoretical analysis of thermal triggers in that the greatest acceleration is observed as the glacier
263 front begins to advance. It is possible that in these latter cases, the thermal activation wave has
264 already reached the terminus by this point and as the glacier forefield is warm, the ice can advance
265 and accelerate unabated (cf. Fowler et al., 2001). Alternatively, if the hydrological system is
266 uniform across the glacier bed, a coincident and glacier-wide switch from efficient to inefficient
267 drainage could explain the monotonic acceleration (Björnsson, 1998).

268 From the twelve velocity datasets we have now derived for Karakoram glacier surge events,
269 there is a mix of evidence relating to the dominant trigger mechanism operating in the region (Table
270 3). A number of characteristics support the surges being thermally rather than hydrologically
271 controlled: (1) the shape of the build-up, active surge and termination phases of the Karakoram
272 surges contrast with those reported from Alaskan glaciers (e.g. Burgess et al., 2012), where
273 hydrology is the surge control. Significantly, in Alaskan glacier surges, the termination phase is

274 much more abrupt than the initiation phase, tending to last several days (or even hours) as opposed
275 to months (or even years) (Kamb et al., 1987). In the Karakoram, on many glaciers the termination
276 phase can last for years (Figure 9), suggesting in these cases the mechanisms operating are
277 fundamentally different to those operating in Alaska. (2) The length of the build-up phase can be of
278 the order of several years in the case of Karakoram surges as opposed to several months as would
279 be predicted by the hydrological surge initiation model. Indeed, Mayer et al., (2011) cited this as the
280 main conflict between their observed and modelled dynamics, suggesting the three-year build-up
281 phase of the Gasherbrum surge greatly exceeded the expected time to switch between an efficient
282 and inefficient drainage system. (3) The timing of the initiation and termination phases appears to
283 be independent of any seasonal control. Hydrologically controlled surges tend to initiate during
284 winter months and terminate during summer months; the Karakoram surge data presented here and
285 elsewhere do not conform to this pattern. (4) Peak velocities are consistently reached during
286 summer months in Karakoram surges. If the surge control was hydrological, we might expect there
287 to be a deceleration during summer months (cf. Kamb et al., 1985) when the basal hydrology would
288 be relatively efficient. (5) There is no evidence of subglacial water either at the margins or within
289 crevasses on the surging glaciers of the Karakoram, observations that have been used elsewhere to
290 support a theory of elevated water pressure being a major control on surging (e.g. Jiskoot et al.,
291 2001). (6) There have been no observations of short-lived, large-scale velocity variations that were
292 a feature of the Variegated Glacier surge and other hydrologically-controlled surges (e.g. Kamb et
293 al., 1985).

294 Intriguingly, however, two main features of the observed Karakoram surges do not conform to
295 thermally-controlled events elsewhere: (1) the return periods of Karakoram glacier surges are
296 significantly shorter than those reported for thermally-controlled surges elsewhere, being of the
297 order of several decades rather than several centuries (Quincey and Luckman, 2014). In all eight
298 cases studied here, the last known surge was pre-1992 (confirmed by the satellite record), so we can
299 report that their return periods are at least 15 years. (2) Karakoram surges tend to last for much

300 shorter periods than those in Svalbard, for example (~3-5 years, as opposed to ~10 years). In
301 extreme cases, they can last as little as 1-2 years, as with the Shakesiga Glacier (Figure 9). This
302 short-lived switch from slow to fast flow resembles Alaskan-type surges more than the Svalbard-
303 type.

304 The dynamics of Karakoram glacier surges do not therefore fit neatly into the well-cited
305 dynamic classification of thermal and hydrologically-controlled surges. There are many remaining
306 unknowns in the Karakoram region that are all likely to play a role in surge magnitude and
307 frequency and may help to explain the inconsistency. The greatest gap in Karakoram glacier
308 knowledge relates to glacier basal conditions, in terms of their thermal characteristics, their
309 composition and their roughness. Previous work has suggested that cold ice may predominate at
310 high-elevations and around the margins of the larger debris-covered glaciers (e.g. Quincey et al.,
311 2009), but based only on seasonal variations in surface velocity. Indeed, given the extreme relief of
312 the Karakoram mountains and the elevation range over which glaciers can be found, it is likely that
313 many different thermal regimes are present, making conventional classes such as warm, cold and
314 polythermal, devised for other contexts, inappropriate for these glaciers (Hewitt, 2014). Similarly,
315 little is known about whether the beds of these surging glaciers are hard or soft, although field
316 observations have identified thicknesses of several metres of basal debris (Owen and Derbyshire,
317 1989) indicating that soft sediment may well underlie at least some of the glaciers in the region, but
318 not necessarily all. Even less is known about their roughness, which may determine the rate of
319 sliding and mass flux if the underlying sediment is immobile (Zoet and Iverson, 2015). Finally, the
320 region is geologically complex, with most surging glaciers crossing two or more major formations
321 (Hewitt, 1998), and possibly underlain by spatially variable geothermal heat flow (Chamberlain et
322 al., 2012).

323 Karakoram glaciers are situated at much higher elevation than those in other surge-prone
324 regions of the world, and are generally shorter and much steeper (Hewitt, 1998). It might be
325 reasonably expected that the overall surge cycle may be much more frequently occurring and

326 shorter lived simply because the accumulation areas of the Karakoram glaciers cannot store vast
327 volumes of ice as can their Polar counterparts. Based on the evidence presented here, we suggest
328 that the thermal, sedimentological and geomorphological characteristics of Karakoram glaciers may
329 vary even on a glacier by glacier basis, and thus the classic thermal and hydrological classification
330 is not appropriate in the Karakoram context. We propose that Karakoram glacier surges have
331 individual surge behaviours, and cannot be collectively characterised. The implication of this is that
332 regional triggers may also not be singularly defined, and are likely to differ even on an individual
333 glacier basis.

334 **6. Conclusions**

335 Using cross-correlation feature tracking applied to optical satellite imagery we have made a
336 significant addition to existing data describing the temporal and spatial evolution of Karakoram
337 glacier surges. These data demonstrate that 1. Karakoram surges are generally short-lived, lasting
338 between 3 and 5 years from initiation to termination, although longer in some cases, 2. The
339 initiation and termination phases are rapid (months to years long) and do not appear to be seasonally
340 controlled, 3. The frontal advances of some small surging glaciers can exceed 1 km over several
341 years of surging, 4. Surge fronts are present in some Karakoram surges, but may simply reflect
342 individual glacier geometries, 5. Uniform acceleration and deceleration across the whole glacier
343 surface, more typically characterises these fast-flow events, 6. Maximum velocities are of the order
344 of 2 km a^{-1} as has been reported in previous work, and 7. Surging tends to peak, and often
345 decelerate, during summer months. Their dynamic evolution does not therefore fit neatly within
346 either of the classically cited thermal or hydrological models of surging, suggesting factors that we
347 still have little knowledge about (e.g. basal thermal and sedimentological conditions) are likely to
348 be dominant controls.

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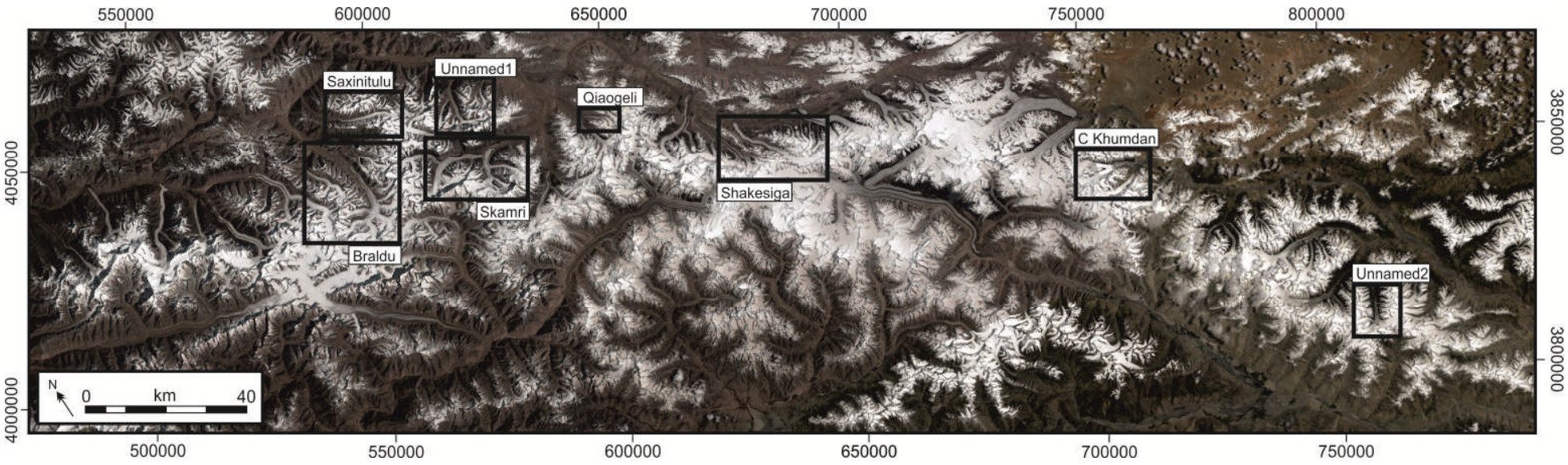
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445 **Acknowledgements**

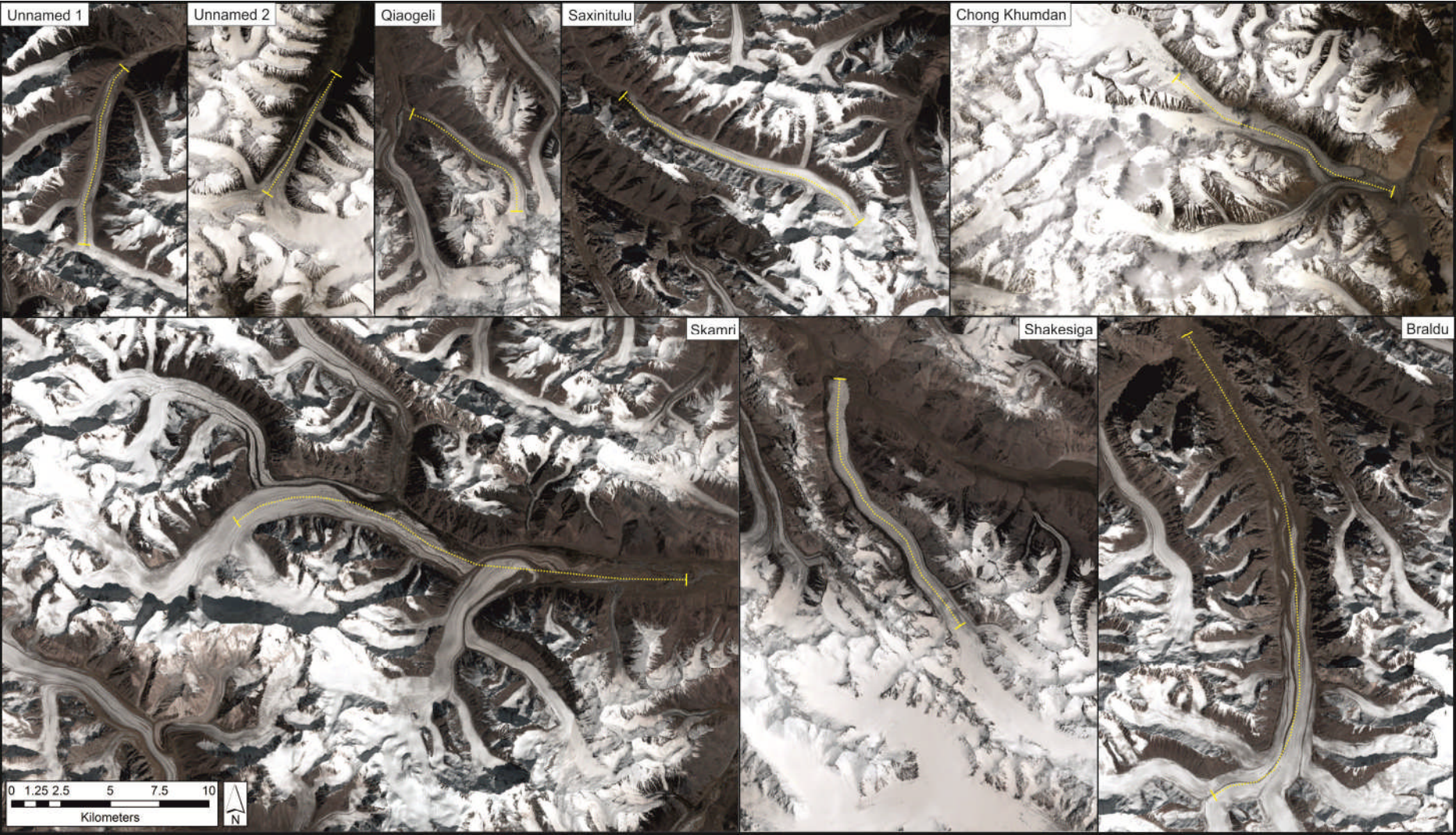
446 The authors would like to acknowledge the provision of ALOS AVNIR imagery through research
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448 All data used in the formulation of this manuscript are available from the first author on request.

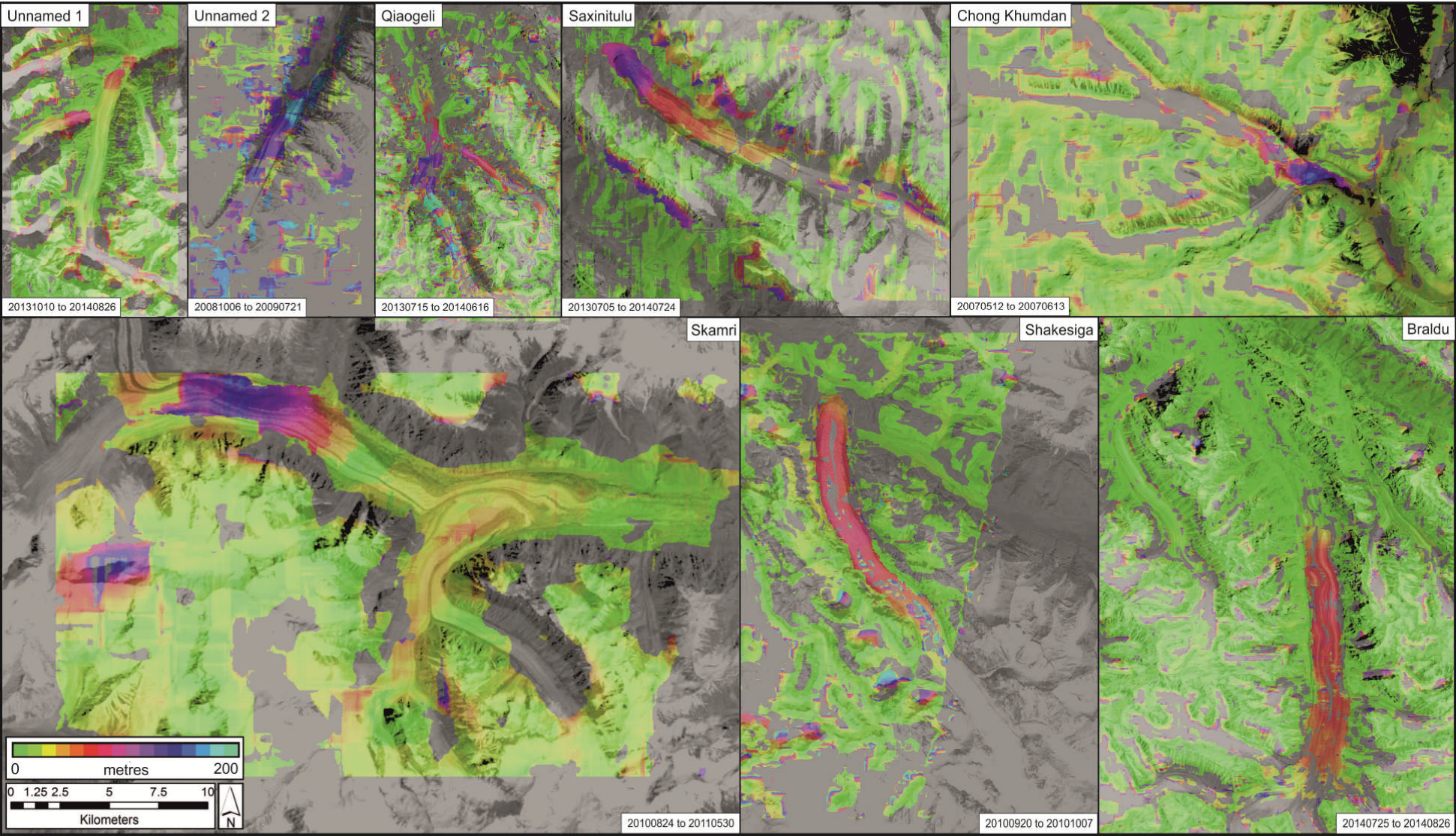
449 Figure 1: The Karakoram region and the location of the eight glaciers analysed in this study. Landsat background imagery © USGS, 2009+2010. Co-
450 ordinates are given in UTM WGS84 Zone 43N. Note the image has been rotated counter-clockwise from true north.



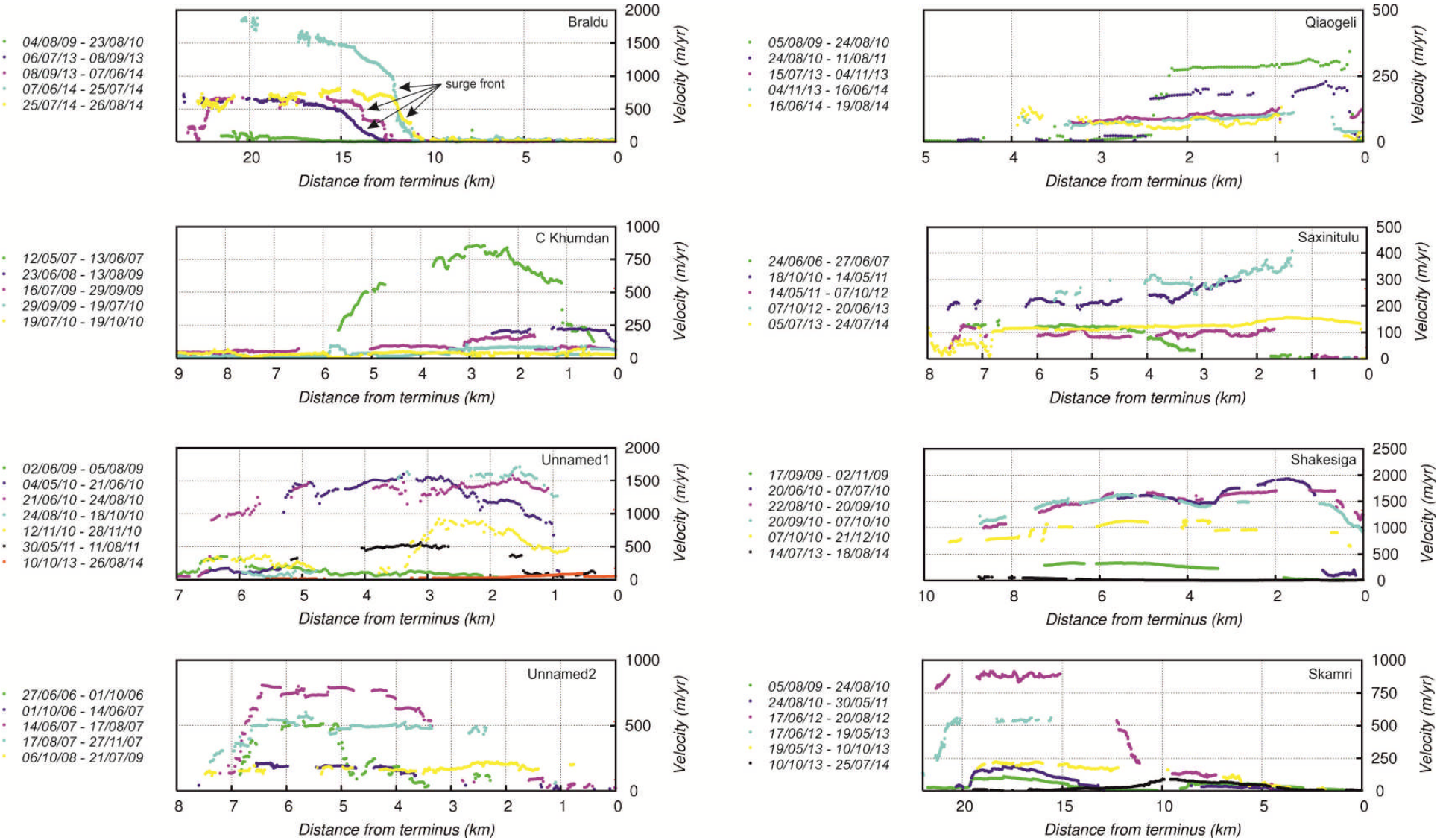
460 Figure 2: Detailed view of the eight glaciers and the centreline profiles used to extract velocity data (shown in Figure 4). In each case the profile is
461 taken from the maximum terminus position reached during each glacier surge, and from the terminus moving up-glacier.



464 Figure 3: Selected filtered velocity fields for each of the eight glaciers.



467 Figure 4: Centreline velocity profiles characterizing the dynamic evolution of surges on each of the eight glaciers in the study. For error estimation see
 468 Table 2. Axes scales are not directly comparable. Note that surge velocities are between one and two orders of magnitude greater than quiescent
 469 velocities in each case, and the clear down-glacier migration of a surge front in the Braldu dataset (labelled).



471 Figure 5: Before and during the surge of Saxinitulu Glacier. The surge began in 2009 and peaked in 2013. The glacier terminus is still advancing in
472 2015 imagery.

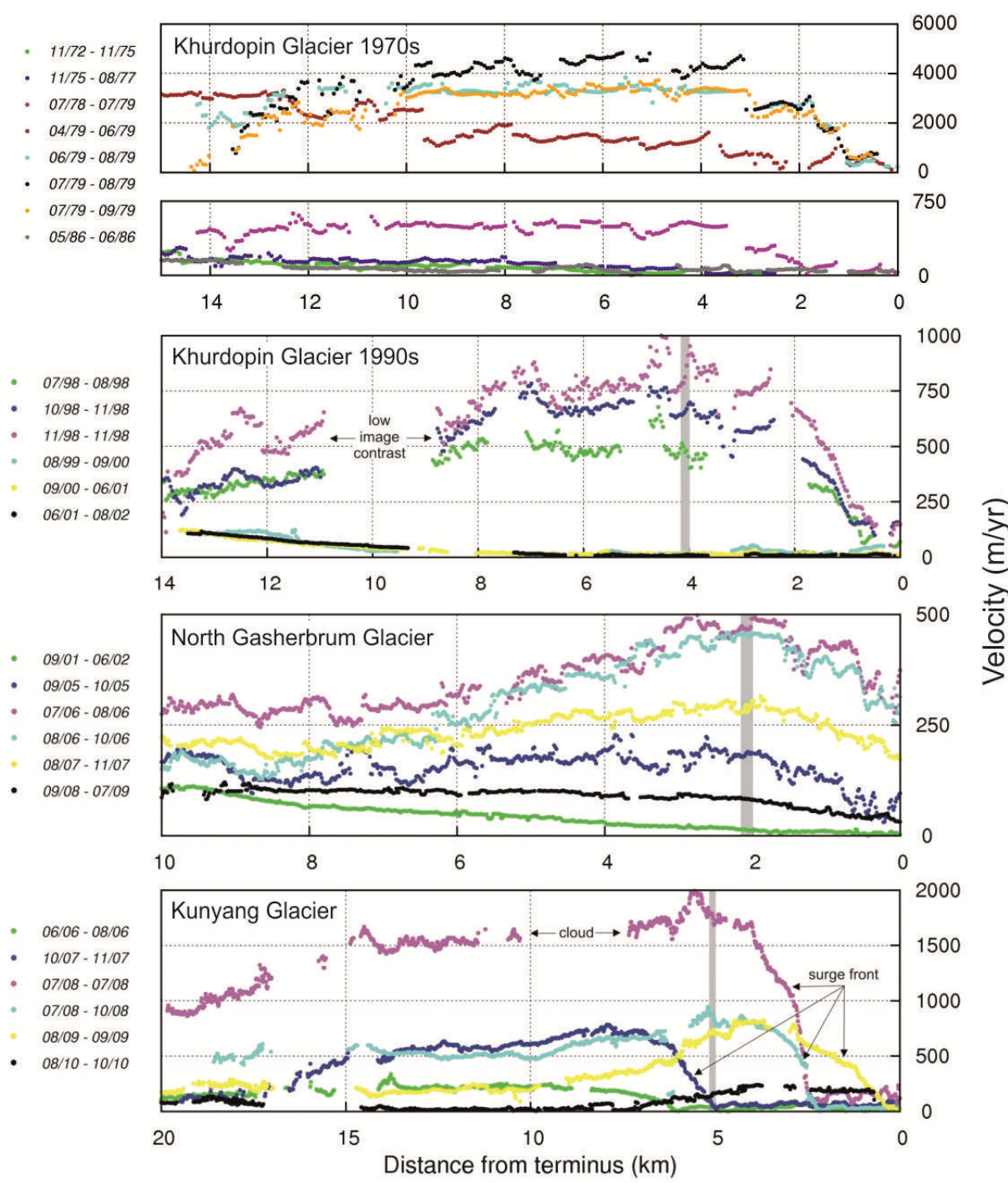
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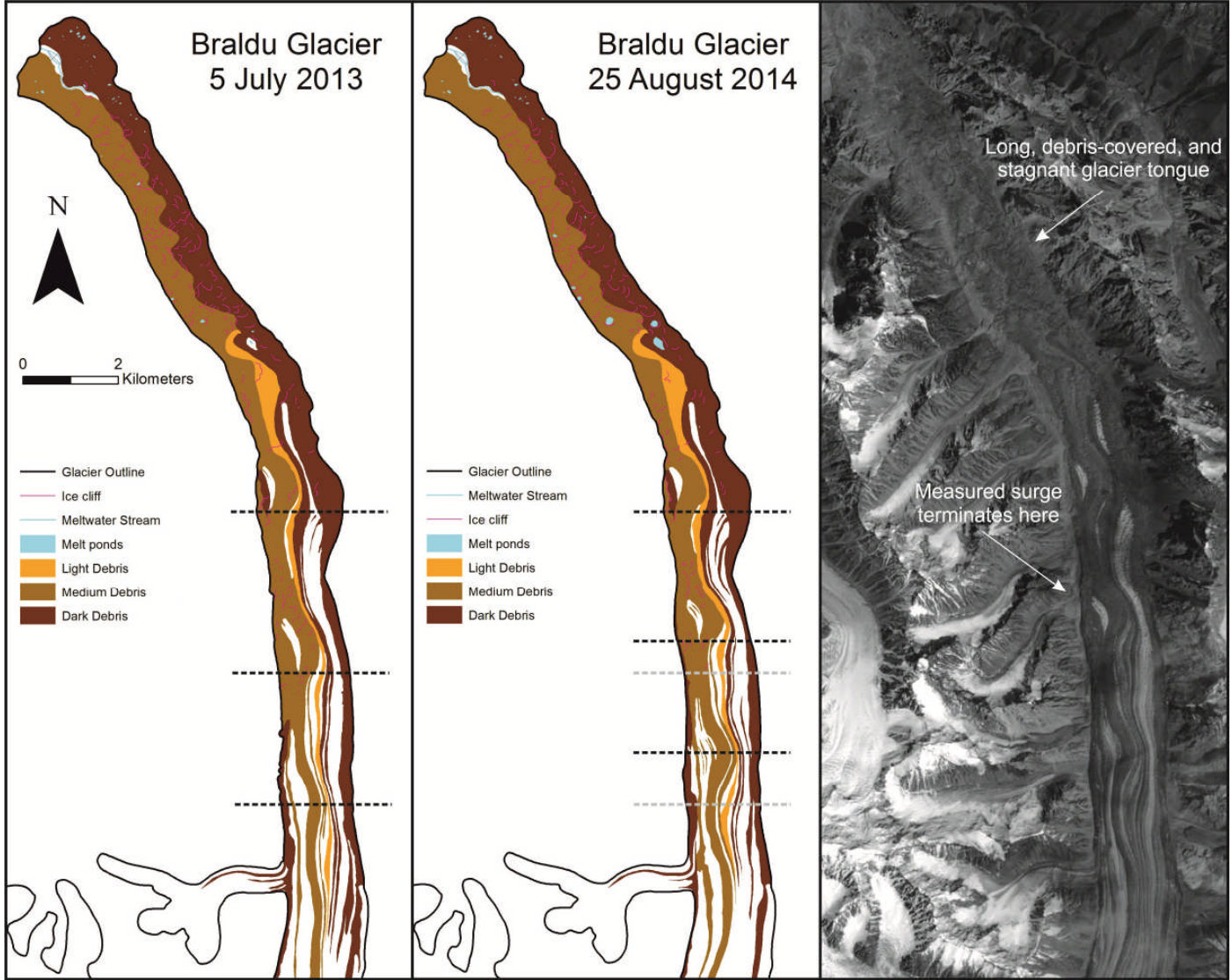
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476 Figure 6: Velocity data for four previously published surges on a) Khurdopin Glacier (during the
477 late 1970s; Quincey and Luckman, 2014), b) Khurdopin Glacier (during the late 1990s), c)
478 Gasherbrum Glacier, and d) Kunyang Glacier (Quincey et al., 2011).

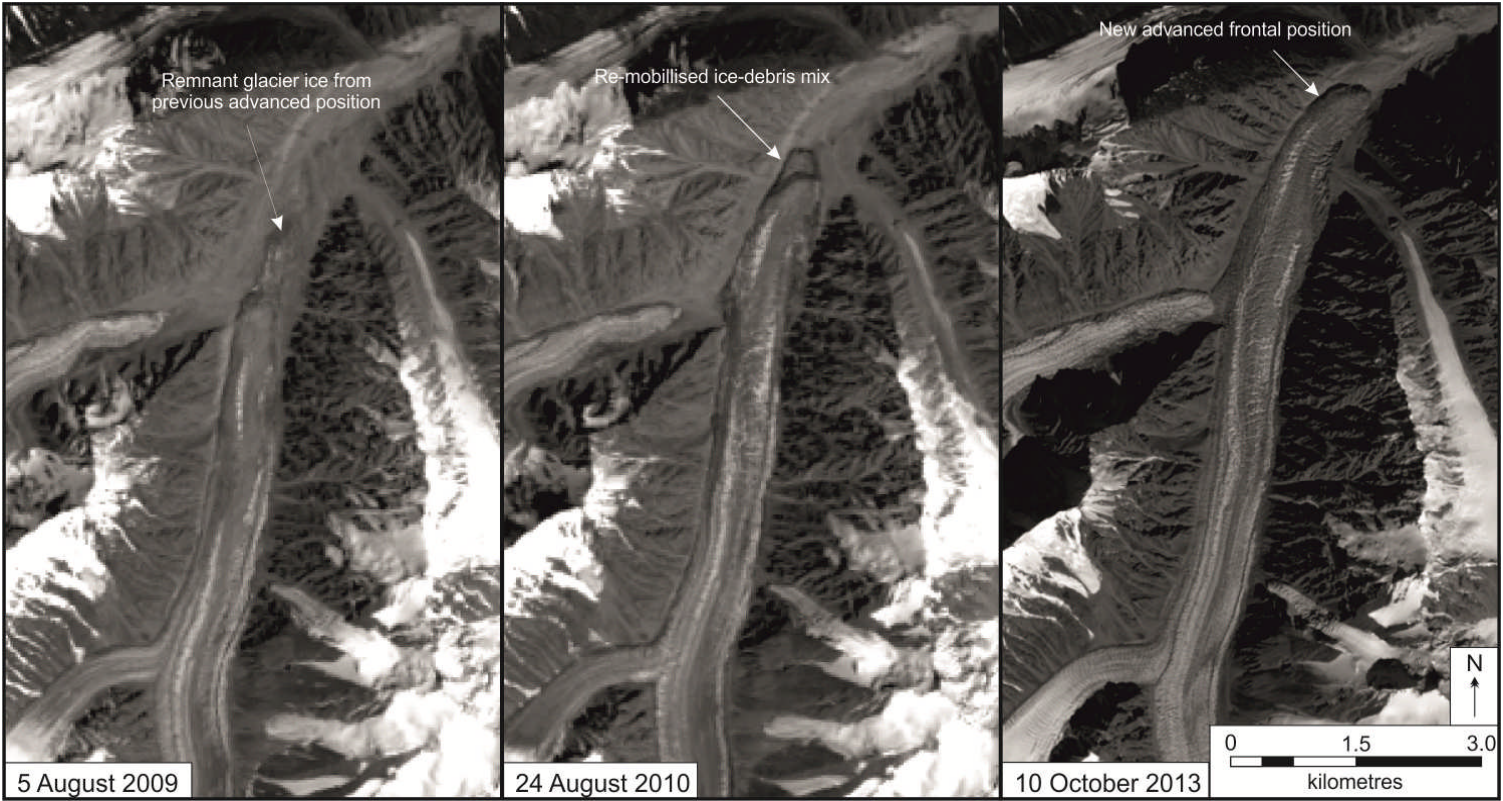


483 Figure 7: The geomorphic context of the Braldu surge. Black dashed lines indicate prominent surface features and their relative positions in each
484 dataset. Grey dashed lines in the August 2014 dataset denote the relative position of each feature in the July 2013 dataset. Note the long debris covered
485 tongue that provides a major obstacle to the down-glacier propagation of the surge front.



487 Figure 8: Evolution of the Unnamed1 surge. Note the former glacier position approximately one kilometre down-valley of the active terminus in 2009,
488 and the way in which that ice-debris mix is overridden by the most recent surge event.

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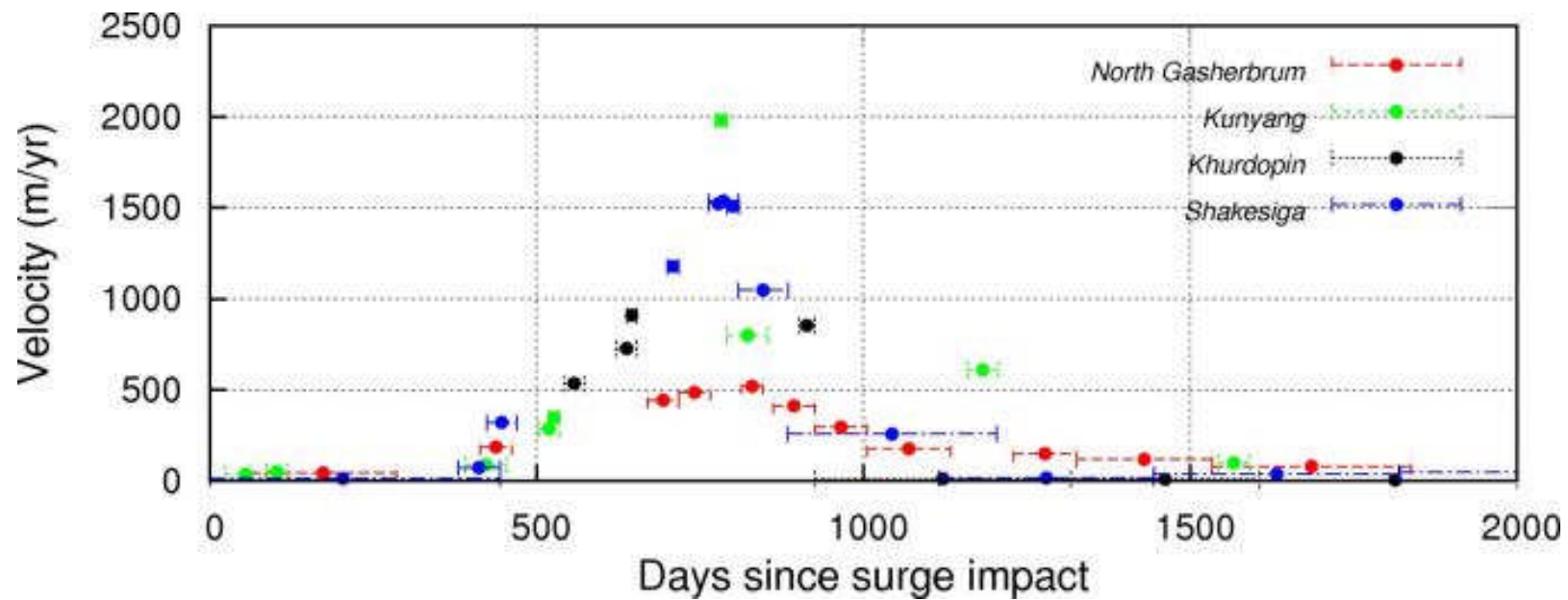
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495 Figure 9: Surge evolution of previously measured events in the Karakoram (Quincey et al., 2011) and the Shakesiga event measured here. Note the
496 shape of the acceleration and deceleration resembles those with a thermal control in Svalbard (Murray et al., 2003), but that the relatively short overall
497 surge period (~600 - 900 days in each case) is more akin to the sudden acceleration and deceleration of hydrologically controlled surges in Alaska
498 (Kamb et al., 1985).

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507 Table 1: Selected characteristics of glaciers in this study (nb. elevations and lengths are approximate values)

Glacier name	Latitude (dec deg)	Longitude (dec deg)	Max elevation (m.a.s.l.)	Min elevation (m.a.s.l.)	Length (km)	Debris covered	Aspect (degs)	Last known surge?	Reference (if applicable)
Braldu	36.143	75.865	6300	3970	34	✓	0	Unknown	Copland et al., 2011
Chong Khumdan	35.183	77.679	6370	4720	20	✓	110	1927-1928	Copland et al., 2011
Qiaogeli	35.967	76.456	7067	4777	9.5	partly	310	1990-2000	Copland et al., 2011
Saxinitulu	36.281	75.943	6286	4600	16.5	✖	290	Unknown	Gardelle et al., 2012
Shakesiga	35.715	76.851	7030	4420	26	✖	320	Unknown	-
Unnamed1	36.178	76.202	6956	4340	14	partly	10	Unknown	-
Unnamed2	34.605	77.978	6435	4746	11	Partly	20	Unknown	-
Skamri	36.055	76.178	6700	3989	40.5	partly	90	1978?	Copland et al., 2009

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509 Table 2: Calculated error in each of the velocity datasets presented

	Images matched	Temporal separation (days)	Sensor	Pixel resolution (m)	Calculated uncertainty (m/yr)
Braldu	20090804 to 20100823	384	TM	30	7
	20130706 to 20130908	64	ETM+	15	21
	20130908 to 20140607	272	ETM+	15	5
	20140607 to 20140725	48	OLI	15	29
	20140725 to 20140826	32	OLI	15	43
Chong Khumdan	20070512 to 20070613	32	ASTER	15	43
	20080623 to 20090813	416	TM	30	7
	20090716 to 20090929	75	AVNIR	10	12
	20090929 to 20100719	293	AVNIR	10	3
	20100719 to 20101019	92	AVNIR	10	10
Qiaogeli	20090805 to 20100824	384	TM	30	7
	20100824 to 20110811	352	TM	30	8
	20130715 to 20131104	112	ETM+	15	12
	20131104 to 20140616	224	ETM+	15	6
	20140616 to 20140819	64	OLI	15	21
Saxinitulu	20060624 to 20070627	368	ASTER	15	4
	20101018 to 20110514	208	ASTER	15	7
	20110514 to 20121007	512	ASTER	15	3
	20121007 to 20130620	256	ASTER	15	5
	20130705 to 20140724	384	ETM+	15	4
Shakesiga	20090917 to 20091102	46	AVNIR	10	20
	20100620 to 20100707	17	AVNIR	10	54
	20100822 to 20100920	29	AVNIR	10	31
	20100920 to 20101007	17	AVNIR	10	54
	20101007 to 20101221	75	AVNIR	10	12
	20130714 to 20140818	400	ETM+	15	3
Unnamed 1	20131010 to 20140826	320	ETM+	15	4
	20090602 to 20090805	64	TM	30	43
	20110530 to 20110811	73	TM	30	38
	20101112 to 20101128	16	TM	30	171
	20100824 to 20101018	55	TM	30	50
	20100621 to 20100824	64	TM	30	43
	20100504 to 20100621	48	TM	30	57
Unnamed 2	20060627 to 20061001	96	ETM+	15	14
	20061001 to 20070614	256	ETM+	15	5
	20070614 to 20070817	64	ETM+	15	21
	20070817 to 20071127	102	ETM+	15	13
	20081006 to 20090721	288	ETM+	15	5
Skamri	20090805 to 20100824	384	TM	30	7
	20100824 to 20110530	279	TM	30	10
	20120617 to 20120820	64	ASTER	15	21
	20120617 to 20130519	336	ASTER	15	4
	20130519 to 20131010	144	ASTER	15	10
	20131010 to 20140725	288	ASTER	15	5

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512 Table 3: Surge characteristics for all twelve events in the Karakoram that have been observed with multi-temporal velocity data. The presence of each
513 characteristic is denoted by • = weak presence to •••• = strong presence; where there is insufficient data to assess the characteristic we state ‘no data’.

Source	Glacier	Surge front	Terminus advance	Winter initiation	Summer termination	Monotonic acceleration	Initiation shorter than termination	Peak velocity in summer
This study	Braldu	••••	No presence	No data	••••	••	No data	••••
	Chong Khumdan	No presence	No presence	No data	No data	No data	No data	•
	West Qogori (Qiaogeli)	No presence	••••	No data	•	•••	No data	No data
	Saxinitulu	No presence	••••	No data	No data	•••	No data	No data
	Shakesiga	No presence	••	••••	No presence	••••	•	••••
	Unnamed1	•••	••••	••••	No presence	•••	••	••••
	Unnamed2	No presence	••••	No presence	••••	•	•	••••
	Skamri	No presence	No presence	No presence	•	••	•	••••
Quincey and Luckman, 2014	Khurdopin (1970s)	No presence	No presence	••	••••	•••	No data	••••
Quincey et al., 2011	Khurdopin (1990s)	No presence	No presence	No data	No data	•••	••	••
	North Gasherbrum	No presence	No presence	•	••	••••	••••	••••
	Kunyang	••••	••••	••	••••	No presence	••••	••••